Natural Ice-Forming Nuclei in Severe Convective Storms

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ABSTRACT

There are two populations of aerosol particles in severe storms: normal background aerosol and aerosolized soil particles. Concentration of the latter, which depends on local wind speed and soil conditions, may be orders of magnitude higher than that of the former. Condensation nuclei are derived principally from the first source. Concentration of ice-forming nuclei, which derive from the soil particles, increases during storms up to 100 times the pre-storm (background) value. This concentration increase is less than that of the aerosol population, indicating that only a fraction of soil particles exhibit ice-nucleating properties. The fraction of soil particles active as ice-forming nuclei in a given particle size range increases with particle size; however, the concentration of ice-forming nuclei in air is counteracted by a decrease in the concentration of aerosol particles with size. Supercooled water drops are nucleated by hydrosol soil particles at temperatures as high as −5.3C.

The quantity of water vapor released during the freezing of supercooled water drops was determined theoretically and experimentally. This value depends primarily on the size of water drops and, to a lesser degree, on the temperature of supercooling. The released water vapor, equal to 0.03 to 3.5 mg per 1–5 mm diameter drops, produces high supersaturation with respect to water at the temperature of the environment in a volume of 300 to 10^5 cm^3, respectively. The water vapor recondenses on cloud droplets and aerosol particles acting as condensation nuclei at higher supersaturation. Some of the aerosol particles acting as ice-forming nuclei will form ice crystals in the water vapor recondensation zone, and these particles will propagate the ice phase within an updraft. Giant aerosol particles, after becoming hydrosol particles, are the most effective freezing nuclei derived from the soil and should be responsible for the appearance of ice at the lowest altitude (warmest temperature).

The freezing temperature spectrum of different hydrosols made of various ices separated from natural hailstones revealed that the warmest freezing temperature was not necessarily associated with hailstone embryos. This indicates that many hailstone embryos form at higher altitudes (lower temperature zones) rather than forming at a freezing level corresponding to the temperature at which the warmest ice-forming nuclei are active.

1. Introduction

In severe storms the ice-forming nucleus is an ice crystal or an aerosol or hydrosol particle which is capable of initiating the formation of the ice phase in the liquid phase of a cloud. An aerosol particle, dry or wet, may collide with a supercooled cloud drop and nucleate the ice phase on contact with its surface. If the supercooled water drop is at a temperature above the ice nucleation temperature of the colliding aerosol particle, the latter may be captured and become a hydrosol particle. The transfer into the liquid phase may also take place during condensation of water vapor on an aerosol particle that acts first as a condensation nucleus. The ice phase will be initiated inside the water drop when supercooling reaches the temperature at which the hydrosol particle is an active ice-forming nucleus. Giant aerosol particles, e.g., 200 μ diameter, will initiate raindrop formation through accretion of cloud droplets. Such a supercooled raindrop upon freezing will release a certain quantity of water vapor

TABLE 1. Concentrations of different groups of aerosol population (number of particles per cubic meter).

Particle diameter (μ)	Aerosol concentration: During no-storm activity, At 1.5 km Average*	During no-storm activity, Ground level Average	During severe storms, Ground level Sustained for minutes	During severe storms, Ground level Maximum recorded
0.26–11	1.1×10^7 (1)			
1.5–5	4.3×10^4 (2)	5.2×10^4 (3)	5×10^5–5×10^6	10^8
5–11	6.1×10^3 (2)	5.0×10^4 (3)	10^5–5×10^5	2×10^6
11–45			5×10^4	5×10^5

* After Blifford and Ringer (1969), averages for 1966.
(1) Year.
(2) May–July.
(3) Average for June–July at New Raymer, Colo.

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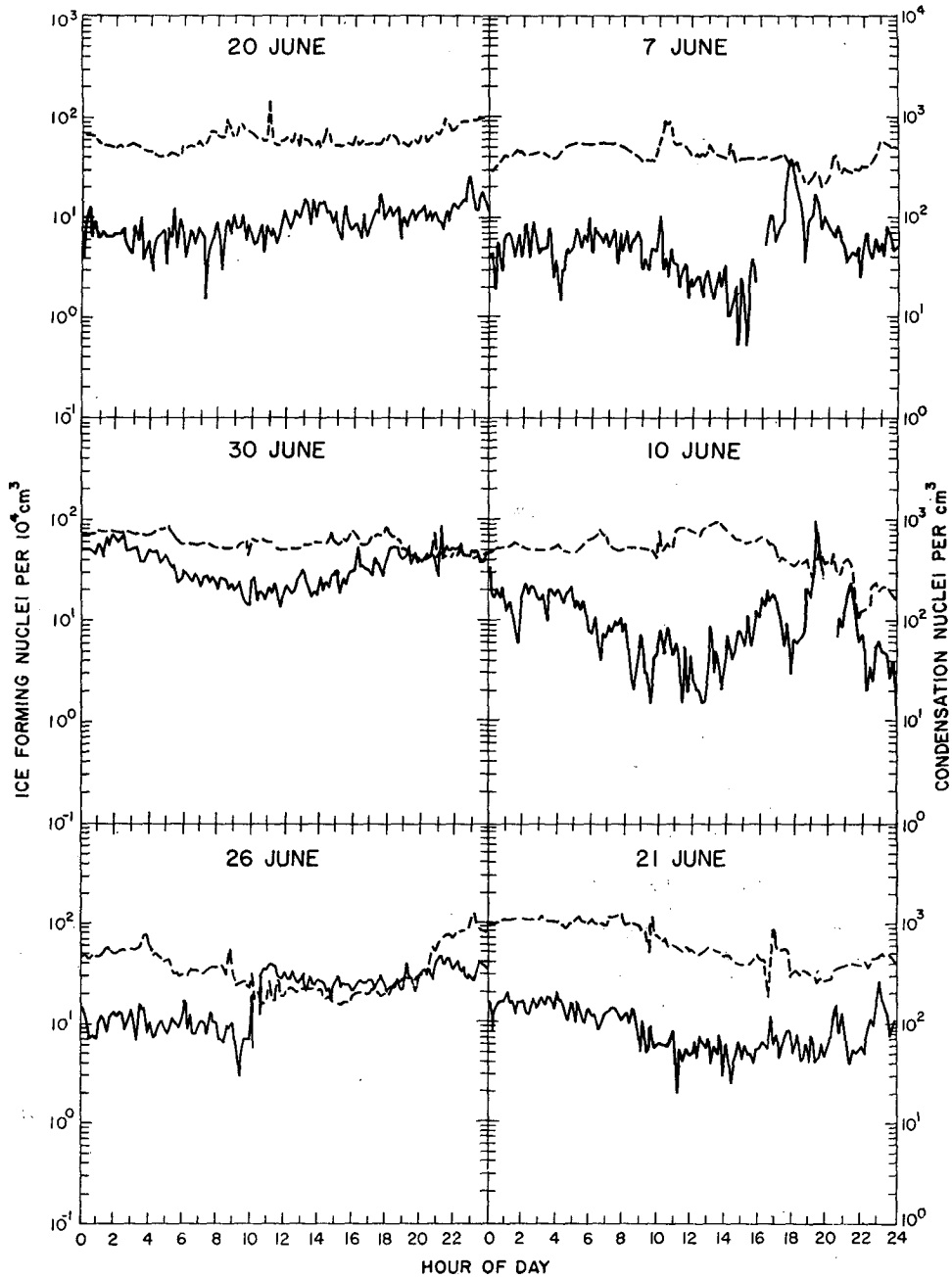

FIG. 1. Concentrations of condensation (dashed lines) and ice-forming (solid lines) nuclei for "no-storm" days (20, 30 and 26 June 1969) and storm days (7, 10 and 21 June 1969).

which subsequently will recondense on aerosol particles present in its wake. These recondensation products may contain ice crystals which should propagate the ice phase within a supercooled cloud.

2. Relationship between ice-forming nuclei and aerosol particles

Average aerosol particle concentrations, by number, for "no-storm" activity days, and maximum concentrations for severe storms in the Colorado-Nebraska region are presented in Table 1. The data for no-storm and storm activity days show the existence of two sources of aerosol particles of entirely different concentrations. The first source is the background aerosol population present on non-storm days in the absence of strong winds. The second source consists of particles raised from the land surface by winds associated with storm activity. The concentration of 1.5–11 μ diameter particles (by microscopic sizing) ingested by a severe storm is 10 to 100 times higher than the average background

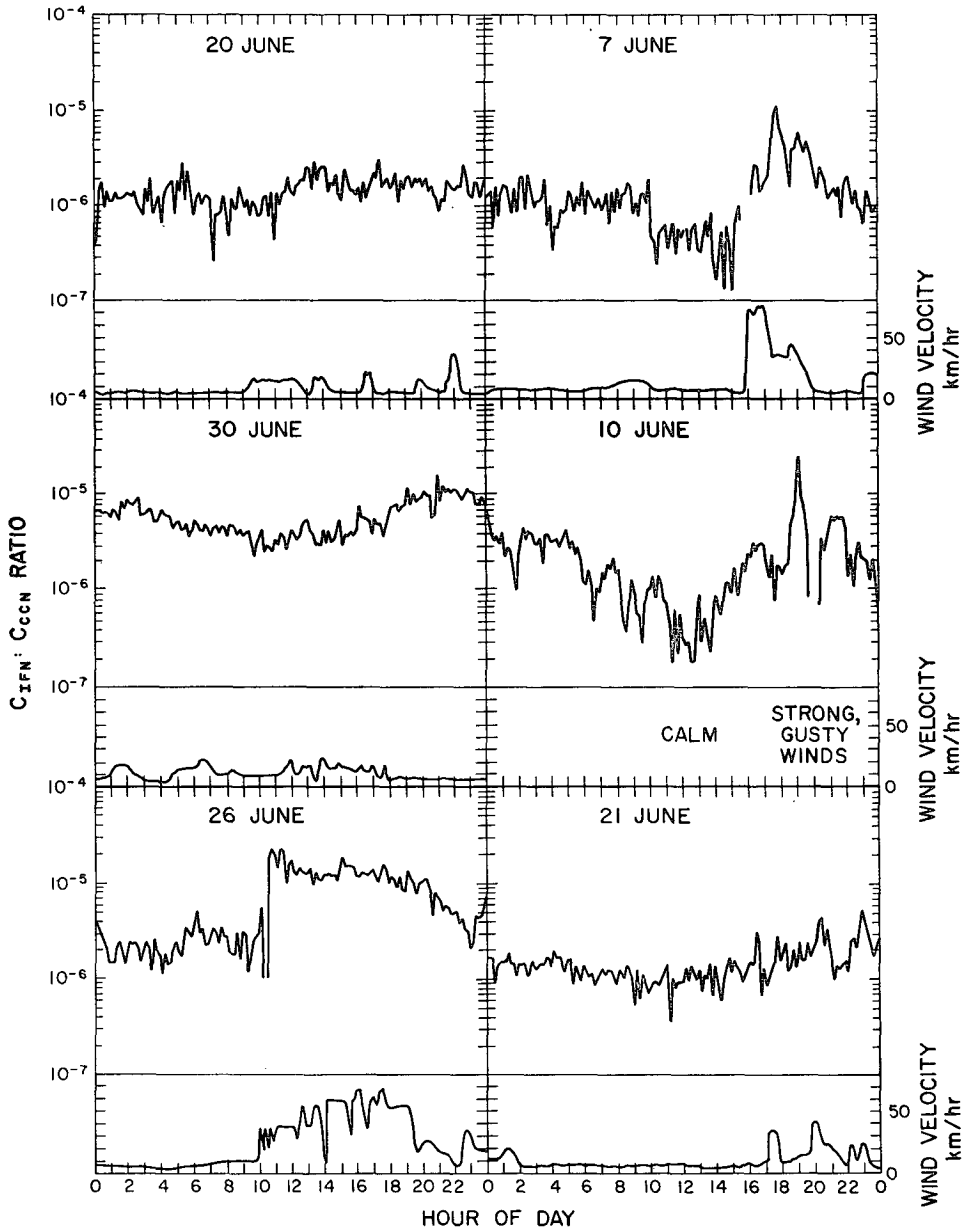

FIG. 2. Ratio of ice-forming nuclei to condensation nuclei and wind velocity data for "no-storm" days (22, 30 and 26 June 1969) and storm days (7, 10 and 21 June 1969).

concentration in this size range. Short-duration pulses as high as 2000 times the background concentration (up to 1.5 gm of soil per cubic meter) were also recorded.

Concentrations of condensation and ice-forming nuclei (−21C) determined by means of the NCAR continuous counters (Langer, 1968; Langer *et al.*, 1967) and their ratios are shown in Figs. 1 and 2 for three "no-storm" days and three days on which severe storms passed the sampling station at various distances. At the very beginning of cumulus formation, the background (aged) aerosol supplies cloud condenstaion nuclei active at very low supersaturations of water vapor. Development of strong convective activity (7 June) or the presence of strong winds (26 June) causes the second source of aerosol particles to predominate (see Table 1). The concentration of cloud condensation nuclei does not show any correlation with large changes of concentration of aerosol particles above 1.5 μ diameter and consequently with wind speed. Changes in the concentration of cloud condensation nuclei seem to be independent of storm activity. If freshly aerosolized soil particles contribute to the condensation nucleus population, then the number must be small.

Ice-forming nuclei quite often belong to the condensation nucleus population (Langer, 1968), and, as shown previously (Rosinski, 1967), they belong to the

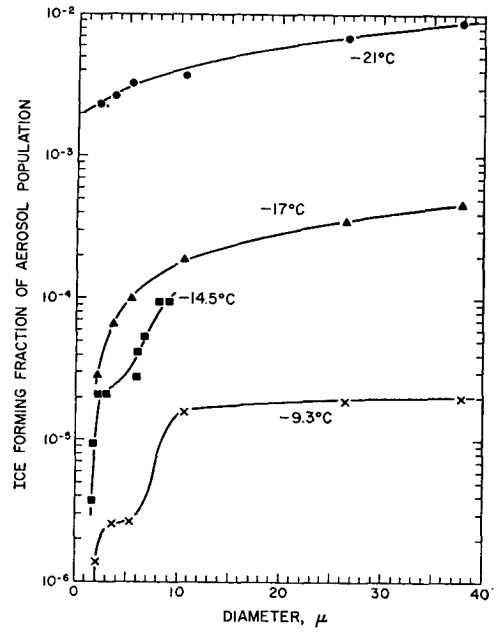

FIG. 3. Cumulative concentration of ice-forming nuclei vs temperature of nucleation.

aerosol population generated by severe storms. The increase in ice-forming nucleus concentration due to a storm was of local character only (7 and 10 June); it did not exist when very severe storms passed 40 mi distant (21 June) from the sampling station. Thus, during storm activity ice-forming nuclei derived from both the background aerosol and particles lifted from local soils by wind action are present; nuclei derived from soil particles are a few to several hundred times more abundant than those derived from the background aerosol. It was frequently observed but has not yet been explained that the concentration of ice-forming nuclei decreased substantially several hours before the onset of storm activity (e.g., 7 and 10 June). Laboratory measurements (Table 2) show that the most effective type of ice nucleation is by contact with large supercooled water drops and by freezing around hydrosol particles inside water drops (80% of the 74–177 μ particles nucleated ice at −10C). The temperature of ice nucleation was determined for soil particles classified in an Andreasen sedimentation pipet. Cumulative concentrations of ice-forming nuclei active at −9.3C, −17C and −21C are presented in Fig. 3.

The size distribution of aerosol particles and ice-forming nuclei derived from soil in the 1.5–15.5 μ diameter size range is presented in Fig. 4. The peak in

TABLE 2. Percent of soil particles acting as ice-forming nuclei in 1.5–11 μ and 74–177 μ water drop diameter size ranges [membrane technique, after Langer (1970)].

Temperature (°C)	Nucleation: By condensation 1.5–11 μ	By condensation 74–177 μ	By contact with: Cloud droplets 1.5–11 μ	By contact with: Large drops 74–177 μ
− 9.3	0.2			
−10.8		7		21
−11.0			0.02	100
−12.0			0.08	
−13.5			0.3	
−16.0			4.3	
−16.2		12		28
−17.0	1.4			
−18.5			10	
−20.7		18		22
−21.0	22			
−22.0			85	

TABLE 3. Percent of particles in each Δd particle size class active as ice-forming nuclei at temperature T.

Δd (μ)	Temperature of nucleation (°C) −9.3	−17	−21
1.5– 2.5	1.0×10^{-4}*	5.0×10^{-3}*	0.1*
2.5– 3.5	3.9×10^{-4}	0.011	0.33
3.5– 4.5	3.6×10^{-4}	0.029	0.52
4.5– 5.5	5.9×10^{-4}	0.063	0.94
5.5– 6.5	2.2×10^{-3}	0.026	1.0
6.5– 7.5	0.023	0.10	1.9
7.5– 8.5	0.041	0.19	1.9
8.5– 9.5	0.043	0.29	3.8
9.5–10.5	0.043	0.30	4.7
10.5–11.5	0.036	0.45	6.8
11.5–12.5	0.018	0.37	7.7
12.5–13.5	0.015	0.57	11.0
13.5–14.5	0.017	0.95	18.0
14.5–15.5	0.026	1.8	26.0

* Estimated.

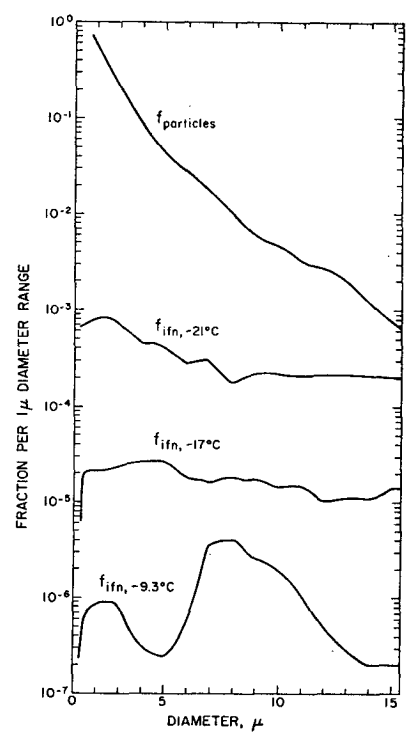

FIG. 4. Fraction of aerosol particles derived from soil in 1.5–15.5 μ diameter size range acting as ice-forming nuclei at different temperatures of nucleation.

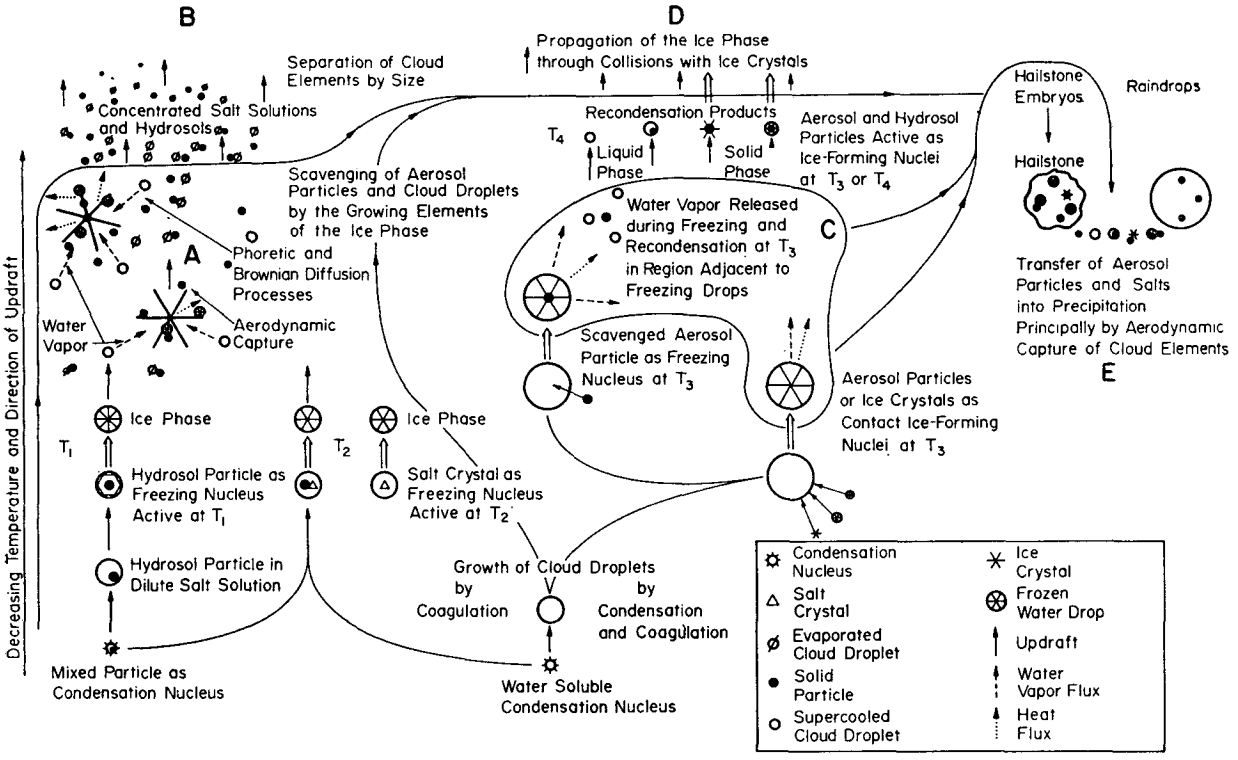

FIG. 5. Schematic diagram of particle interaction in a cloud containing liquid and solid phases. (No relative magnitudes of T_1, T_2, T_3 and T_4 are implied.)

concentration of ice-forming nuclei active at −9.3C exists in the 7.5–11.5 μ diameter size range. This peak, which may be due to the presence of particles of some different origin, corresponds to a concentration peak in the same size range found in some milky hailstone embryos (Rosinski and Kerrigan, 1969). This would indicate that ice-forming nuclei active at higher temperatures nucleate some of the hail embryos in severe storms. The percent of ice-forming nuclei active at different temperatures in different size fractions of soil particles is given Table 3. The increased ability of larger particles to act as ice-nucleants is counteracted by their lower concentrations (Fig. 4).

3. Ice-forming nuclei in precipitation

Different particle-to-hydrometeor transfer processes operate within a severe storm (Fig. 5). A solid particle present in a hailstone or a raindrop may have a warmer ice-nucleating temperature than the rest of the particles in the hydrometeor, yet may not necessarily have initiated its formation. It may be transferred to a hydrometeor by aerodynamic capture (Fig. 5E). Evaporation of cloud droplets will take place in the presence of the ice phase (Figs. 5A, 5B) leading to the formation of cloud particles containing highly concentrated salt solutions. Some of the aerosol particles taking part in the recondensation process around a freezing drop may nucleate the ice phase (Fig. 5C). One freezing drop may, therefore, produce a number of ice crystals and start a chain reaction that propagates the ice phase with an updraft (Fig. 5D). In the following sections an attempt will be made to reconstruct the role which a particle present in a hydrometeor played in the formation of that precipitation.

a. Liquid phase: Rainwater

Rainwater contains varying quantities of organic and inorganic particulate matter (Rosinski, 1967; Rosinski and Kerrigan, 1969) and water-soluble salts (Turner, 1955; Woodcock and Blanchard, 1955) derived from scavenged aerosol particles. It also contains dissolved salts from partially soluble soil particles and water-soluble organic material. The mechanism of the formation of precipitation and its life history have a decisive influence on the relationship between raindrop size and quantities and nature of different sized particles and soluble material. Also, rain may contain melted hail and, therefore, salts and different sized solid aerosol particles transferred into hailstones during their life with a storm. A water drop freezing technique (Bigg, 1953; Vali, 1968) was used to derive temperature spectra of drops made of water from different hydrometeors.

1) STORAGE TIME AND MATERIAL OF CONTAINER. Particle-free distilled water was stored for one year in pyrex glass and polyethylene containers. The spread in temperature of freezing of drops was as follows:

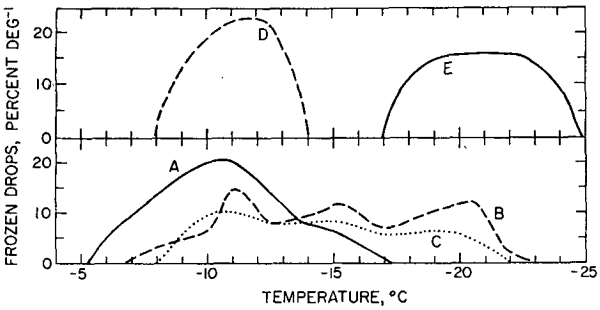

FIG. 6. Freezing temperature spectra of two soil suspensions (A, B, C: New Raymer, Colo.; D, E: Fort Collins, Colo.) filtered through different diameter pore filters: A, 0.22 μ; B, 0.01 μ; C, 0.8 and 0.01 μ, successively; D, 0.8 μ; and E, 0.8 and 0.01 μ, successively.

original water, −25 to −35C; in polyethylene, −26 to −30C; pyrex glass, −16 to −23C. The shift to warmer temperature of nucleation in glass containers was 9C, indicating that the contact of rainwater with the collector may have a decisive influence on the freezing temperature (Wylie, 1953).

2) SOLUBLE FRACTION OF SOIL. A sample of New Raymer, Colo., soil contained 4.4% unbound water and 0.05% soluble material (at 20C). The results of drop freezing of the saturated soil solution/hydrosol filtered through different diameter pore filters are given as spectra A–C of Fig. 6. When diluted (1:5) the solutions gave similar results. Freezing temperature spectra of solution/hydrosol of a sample of a Fort Collins, Colo., soil are presented as spectra D and E. Soil particles smaller than 0.01 μ diameter did not nucleate ice at warmer temperatures in that solution/hydrosol (curve E). These two examples show that suspensions of different soils may have a different mode of ice nucleation.

3) PARTICLES FROM BACKGROUND AEROSOL AS ICE-FORMING NUCLEI. Background aerosol was sampled by means of a sequential sampler in New Raymer on 15 July and 8 August 1969. The warmest freezing temperature of −10C was recorded in one of the samples, indicating that background aerosol particles neither supplied warm freezing nuclei nor formed solutions freezing at temperatures warmer than −10C.

4) SOIL PARTICLES AS ICE-FORMING NUCLEI. A dispersion of soil particles that are difficult to wet in water can be created with the help of a wetting agent. It was found experimentally that a drop of sulfonated dioctyl sodium sulfossuccinate solution when used in concentrations below 10^{-4} gm cm^{-3} of water did not modify the freezing temperatures of soil suspensions. The warmest freezing temperature of hydrosols formed from different soils was −5.3C (for soil from the New Raymer area). Some of the soil suspensions started to nucleate at −14C.

5) RAINWATER. Different samples of rainwater collected in a single storm started to freeze at different temperatures. The warmest freezing temperature (−5C) was associated with the presence of a high concentration of hydrosols. High accumulation of particulate matter was observed in hailstones, and rain samples with high concentration of particles might contain melted hailstones. This is probably the only conclusion which can be drawn from the freezing drop technique.

b. Solid phase: Hailstones

Hydrosol particles are transferred with the advance of the ice front during the phase transition (Uhlmann *et al.*, 1964; Hoekstra and Miller, 1965); this produces a non-uniform concentration of particles in the frozen hydrosol. The same process takes place during the formation of a hailstone, and the interpretation of its growth on the basis of freezing nuclei distribution in different ices within the hailstone must take the migration of particles into account.

A suspension of up to 1.5 μ diameter soil particles in water containing the water soluble fraction of soil was placed in a polyethylene cylinder and was frozen from the outside toward the center. The concentric layers of ice were melted, and the corresponding freezing temperature spectra determined (Fig. 7). Curve B is similar to the temperature spectrum of the original suspension; but the outside ice layer C and the central portion of the cylinder A show enormous changes due to displacement of the dissolved salts and particles relative to the moving interface during freezing. The warmest temperature of refreezing of ices present in a hailstone formed through a wet growth is unrelated, therefore, to the nucleation temperature of particles present in those ices (it should be related to particles present in boundary layers). The refreezing temperature of ice formed during dry growth should furnish information about freezing nuclei present in that portion of a cloud in which this ice grew.

Hydrosol suspensions made from melted (entire) hailstones usually started freezing between −5 and −6C, with some rare exceptions when the temperature was between −11 and −14C. Results of such analyses have little meaning because they do not give any information about the spatial distribution of ice-forming nuclei active at different temperatures within hailstones. In a more exact analysis, hailstones collected from different hailstorms were sliced with a microtome, different ices from a slice melted, and the onset tem-

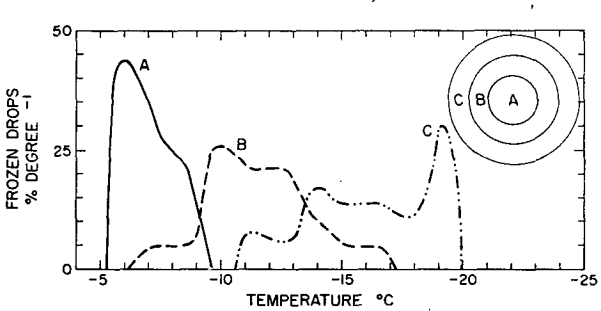

FIG. 7. Temperature spectra of different parts of ice of frozen hydrosol (freezing proceeding from the wall of cylinder toward the center).

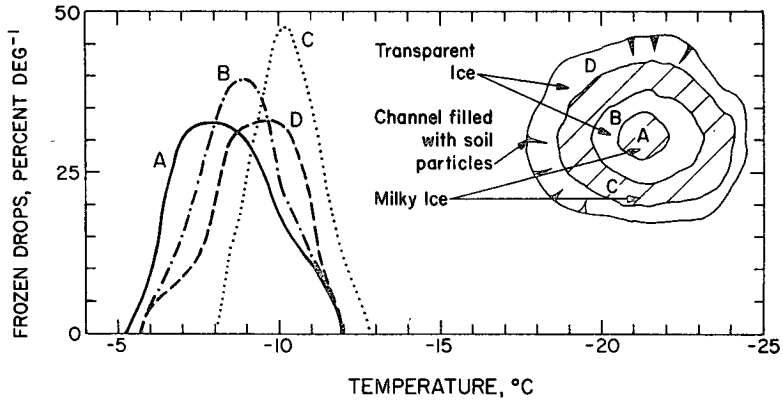

FIG. 8. Temperature spectra of different layers of ice of a hailstone (from Niwot, Colo., 1969).

peratures of freezing of hydrosols were determined. Selected examples are presented below.

1) A 2.5-cm diameter hailstone from a violent storm of short duration (1969, Boulder-Niwot, Colo.) was collected from the ground (Fig. 8). Channels in the outer layer of transparent ice (D) were filled with soil particles. From the temperature spectra of different layers, it seems that the layer of milky ice C (−8C) protected the inner layer B and the embryo A from the rainwater-soil particle suspension from the ground. The freezing temperature of the embryo was −5.3C and was the warmest temperature of all parts of the hailstone.

2) Freezing temperature spectra of hydrosols plotted for different parts of a 3.4 cm-hailstone (1 July 1967, Boulder) are given in Fig. 9. The milky center A and an intermediate milky layer C started to freeze at −12C. The hydrosol from the outer layer D (transparent ice) started to freeze at −7.7C and from the intermediate transparent layer B at −8C. The milky embryo A was formed in the absence of warmer temperature freezing nuclei. The growing hailstone seems to return to the −12C temperature zone; it grew in the presence of a high concentration of particles (curve C). The final growth produced transparent ice with large particles, and the onset temperature of freezing of this solution/hydrosol system was −6.6C. Similar hailstones were found in Nebraska and Northern Colorado hailstorms.

Milky hailstone embryos seem to form by two processes. Some embryos are built preferentially through accretion of supercooled cloud droplets as indicated by the low concentrations in embryos of 1.5–3 μ diameter hydrosol particles; this could result from a "wet" growth of the hailstone embryo. Other embryos are formed by a truly "dry" growth with a large contribution of ice crystals. Transparent ice in the outer layer of a hailstone always contained the warmest ice-forming nuclei which were probably transferred into the ice at lower altitudes during final stages of growth. The majority of hailstones in some storms form around the milky center, with freezing temperatures of different ices decreasing toward the center of the stone; this strongly suggests that the central part, an embryo, did

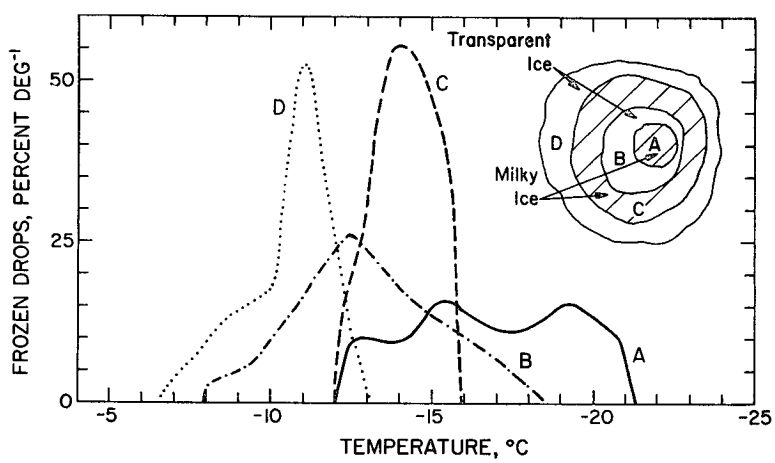

FIG. 9. Temperature spectra of different parts of a hailstone (from Boulder, Colo., 1 July 1967).

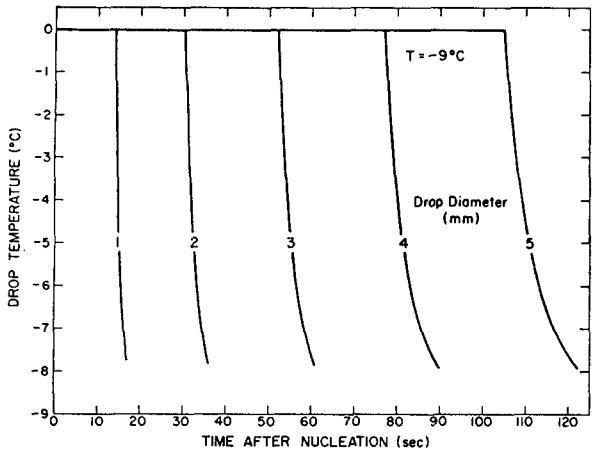

FIG. 10. Drop temperature vs time after nucleation for various drop diameters and an air (supercooling) temperature of −9C.

not originate on the warmest freezing nucleus but was formed around nuclei active at lower temperatures.

4. Formation of the ice phase through evaporation-recondensation of water vapor

The lifetime of giant aerosol particles in a severe storm is extremely short because they coagulate with cloud droplets upon entry into the cloud (Rosinski and Kerrigan, 1969). These particles form large water drops around them, thus becoming hydrosol particles which can act as freezing nuclei. Upon reaching the temperature of nucleation, the large drop freezes, releasing a certain amount of water vapor (Fig. 5C) during the process. As the drop seeks thermodynamic equilibrium, water vapor lost to the atmosphere creates a condition

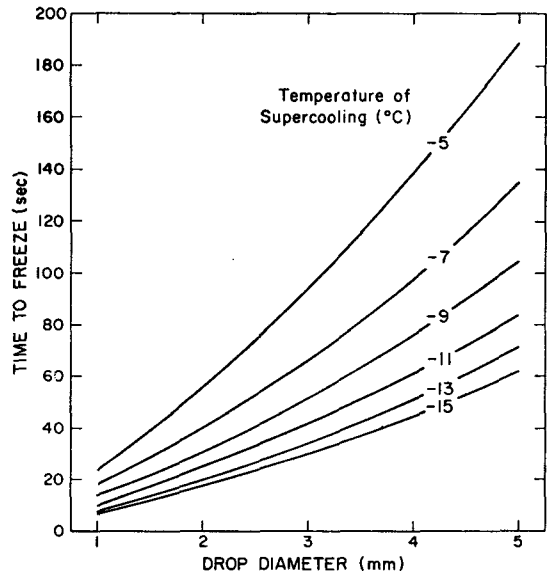

FIG. 11. Time after nucleation for complete drop freezing vs drop diameter for various values of air (supercooling) temperature.

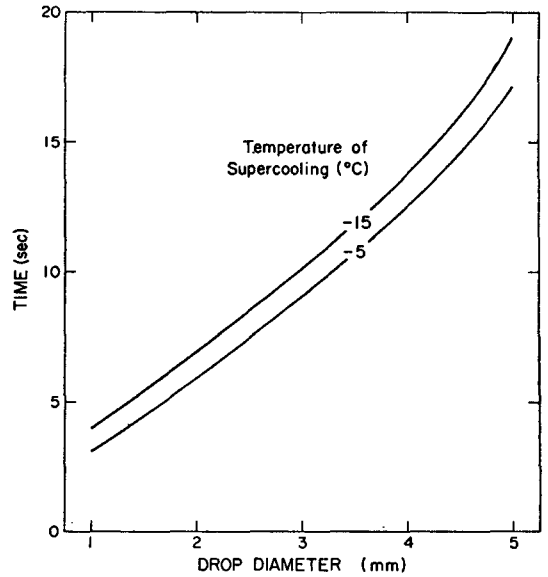

FIG. 12. Time after complete freezing for a drop to cool vs drop diameter for two values of air (supercooling) temperature.

of supersaturation in the drop's wake (Dye and Hobbs, 1968), in which aerosol particles of different kind and size are present.

Heat is supplied to the surface of a drop falling in air by conduction and condensation. The rate of transfer is given by Dye (1967) as

$$\frac{dQ}{dt}=4\pi aK(T-T_a)f+4\pi aLD(\rho-\rho_a)f, \qquad (1)$$

where a is the drop radius [cm]; K the thermal conductivity of air [5.63×10^{-5} cal cm^{-1} sec^{-1} (°C)$^{-1}$]; T the environment temperature [°C]; T_a the sphere surface temperature [°C]; f the ventilation factor [dimensionless] for smooth spheres equal to $1+0.31\,\mathrm{Re}^{\frac{1}{2}}$ (Johnson, 1960); L the heat of vaporization [678 cal gm^{-1} at 0C]; D the diffusion coefficient of water vapor in air [0.219 cm^2 sec^{-1}]; ρ the environment water vapor density [gm cm^{-3}]; ρ_a the water vapor density at sphere surface [gm cm^{-3}]; and Re the Reynolds number [dimensionless].

The rate at which heat is removed at the drop surface equals the rate at which heat is produced in the drop by freezing; thus,

$$L_f\frac{dM}{dt}=4\pi aK(T_a-T)f+4\pi aLD(\rho_a-\rho)f, \qquad (2)$$

where L_f is the latent heat of freezing [79 cal gm^{-1}] and M is the mass of drop present in the liquid phase [gm].

The drop mass lost by evaporation during the freezing period is (Byers, 1965)

$$\frac{da}{dt}=\frac{Df}{a\rho_l}(\rho_a-\rho), \qquad (3)$$

where ρ_l is the density of water [gm cm^{-3}] and t time [sec].

The time rate of change of the sphere's temperature is

$$\frac{dT_a}{dt}=-\frac{3f}{a^2\rho_l}\left[\frac{K}{c_p}(T_a-T)+D\left(\frac{L}{c_p}-T_a\right)(\rho_a-\rho)\right], \quad (4)$$

where c_p is the specific heat of ice [0.505 cal gm^{-1} (°C)$^{-1}$]; ρ_a is taken to be the saturation vapor density over ice at T_a, and ρ the saturation density over water at T (Dorsey, 1940).

Fig. 10 gives a family of cooling curves representing the temperature histories of various diameter drops supercooled to −9C before nucleation. The significant period for vaporization is while the drop is at 0C, and is represented by the horizontal bar.

Fig. 11 gives a family of curves representing the time to freeze for various values of supercooling and drop diameter. Fig. 12 gives two curves representing the time to cool of completely frozen drops to the temperature of environment (supercooling), while Fig. 13 represents the difference in vapor pressure over the surface of the drop at 0C and the ambient vapor pressure at the indicated temperature of supercooling. Fig. 14 gives the total mass vaporized as a function of supercooling for various values of the diameter.

In order to estimate the supersaturation in the volume affected by the evaporating drop, we take the affected volume V to be

$$V=tv\pi d^2, \quad (6)$$

where t is the time for drop to freeze [sec], v the terminal velocity of drop [cm sec^{-1}], and d the drop diameter [cm].

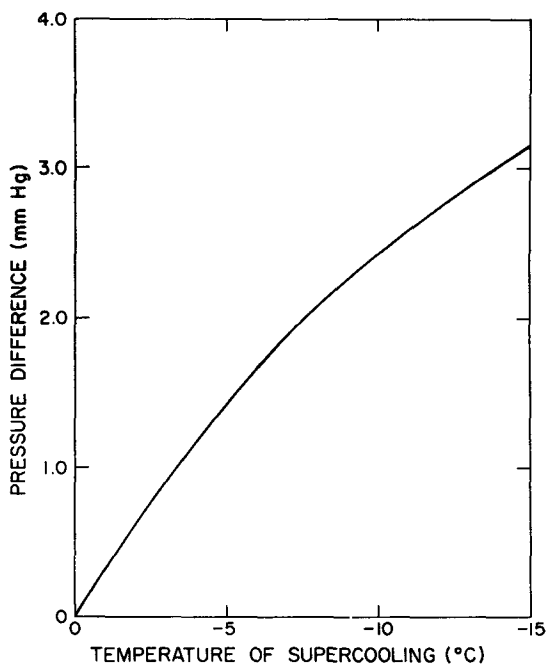

FIG. 13. Water vapor saturation pressure difference over water between 0C and air (supercooling) temperature.

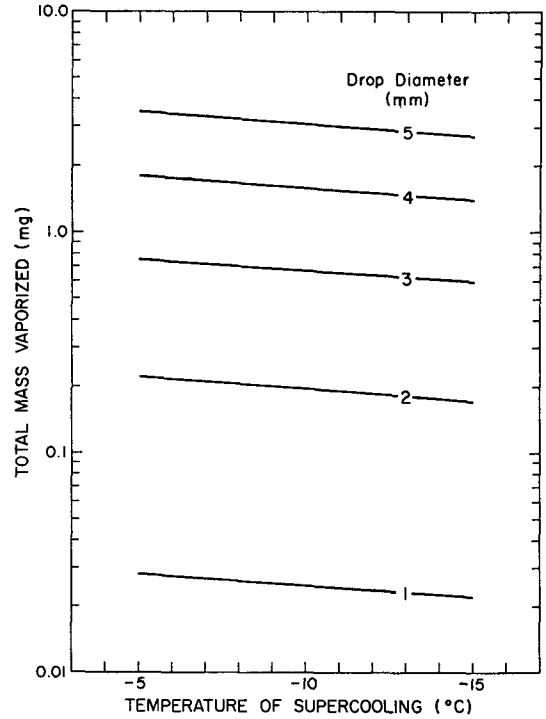

FIG. 14. Total mass of water vaporized from a drop during freezing and cooling vs the temperature of supercooling for various drop diameters.

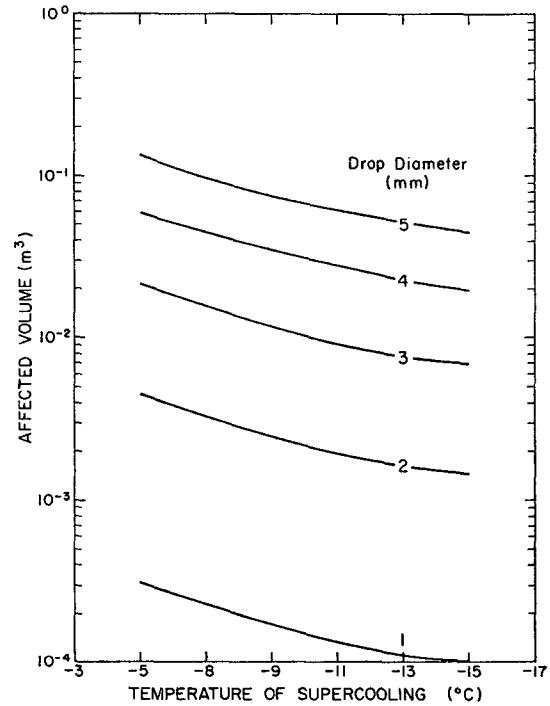

FIG. 15. Estimated volume of air supersaturated by drop water vapor vs temperature of supercooling for various drop diameters.

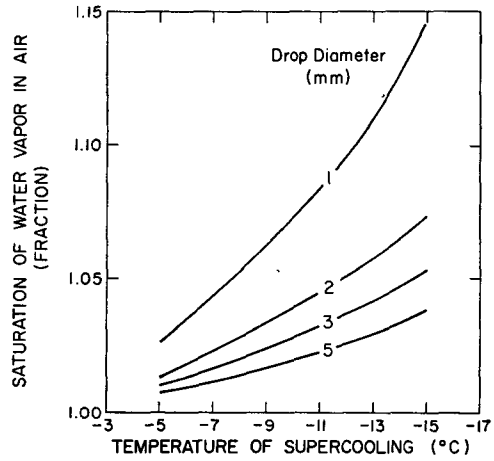

FIG. 16. Estimated mean water vapor saturation produced by drop water vapor vs temperature of supercooling for various drop diameters.

Fig. 15 gives curves showing affected volume vs temperature of supercooling for various values of drop diameter. The water vapor saturation in the form of a fraction is then computed by

$$S=\frac{(W/V)+\rho}{\rho}, \tag{7}$$

where W is the total mass of water vaporized by a given drop [gm]. In each case the supersaturation, $S-1$, is considerable (Fig. 16).

During accretion by a freezing drop, cloud droplets encounter either a liquid or an ice surface, both of which are at ~0C. The mass accreted requires less energy to raise its temperature to 0C than it gives off when it freezes. This effect prolongs the time the drop takes to freeze and increases the affected volume V, although probably not the degree of supersaturation. The heat from a freezing drop supplied to the wake will raise its temperature above the temperature of the environment (for a 3-mm freezing drop at −10C the temperature of the wake will be raised to −9.7C), but this has only negligible effect on recondensation of water vapor on a variety of aerosol particles present within the envelope of recondensing vapor. Recondensation will take place before diffusion could lower the supersaturation to ambient. Laboratory experiments with freezing drops produced ice crystals through the evaporation-recondensation process in the presence of soil particles. Satisfactory agreement was found between the calculated and the experimentally determined masses of water vapor released during water drop freezing. A hail embryo formed through agglomeration of the ice crystals and supercooled cloud droplets above the level of the freezing zone of large water drops should not contain warmer temperature freezing nuclei. The observed high concentration of 1.5–3 μ diameter particles associated with some embryos (Rosinski, 1967) may be due to the rapid coagulation of recondensation products on those particles. It should be noted that the growth rate of an ice crystal growing in an environment at water saturation has a maximum at a temperature around −14C which is close to observed temperatures of ice nucleation of particles present in many of the milky embryos. Under such circumstances ice crystals should grow rapidly to larger sizes, assuming that they are at the temperature of the environment; this, in turn, should enhance the coagulation of cloud particles.

5. Effects of lightning on formation of the ice phase

One of the effects of lightning is the formation of a volume in which cloud droplets should partially or completely evaporate. The evaporated water vapor should therefore be available for recondensation on supercooled droplets and on aerosol particles. Sansom (1966) has shown that in Kenya only 37% of the storms with frequent lightning, as compared to 68% of the the storms with infrequent lightning and 79% of the storms with no lightning, were classed as damaging hailstorms.

Sansom's statistical analysis was performed on the suggestion of Goyer (1965) whose experimental work has shown that shock waves can possibly lead to hailstones' softening. But to be softened by the shock wave, hailstones must already exist within the shock wave channel. One may speculate that propagation of the ice phase within clouds by means of evaporation-recondensation processes leading to redistribution of available liquid water over numerous hail embryos may explain Sansom's correlation more logically.

6. Conclusions

The following general conclusions can be drawn from this study of ice-forming nuclei present in the atmosphere during severe hailstorms:

1) Two sources of ice-forming and freezing nuclei exist in the atmosphere. The first source is derived from the aged background aerosol present in the atmosphere, and the second is produced by the storm itself. This latter source consists of local soil particles aerosolized by the turbulence generated by winds. Its concentration, which may be orders of magnitude higher than that of the first source, depends on the wind velocity and soil conditions. The fraction in a given size range of aerosol particles acting as ice-forming nuclei increases with the size of the particles. The concentration of ice-forming nuclei in air is counteracted by a decrease in concentration of aerosol particles with size.

2) The warmest temperature of freezing of hydrosols made by melting different ice layers of hailstones was found to be not necessarily associated with hailstone embryos. This indicates that many hailstone embryos composed of milky ice form at higher altitudes (lower temperatures) and not at freezing levels corresponding

to the temperature at which the warmest ice-forming nuclei are active.

3) Water drops release water vapor into their surroundings during freezing. The warmest temperature of freezing takes place with those drops formed around giant soil particles. The released water vapor which is supersatured with respect to supercooled water at the temperature of the environment recondenses on cloud particles (aerosol particles and cloud droplets). Some of the aerosol particles may act as ice-forming nuclei and form ice crystals which propagate the ice phase in the direction of the updraft.

The conclusions drawn in this article are based on data collected in the Colorado-Nebraska region, and do not purport to apply generally to all hail-producing storms.

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